The energetics of ocean heat transport

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Abstract

A number of recent papers have argued that the mechanical energy budget of the ocean places constraints on how the thermohaline circulation is driven. These papers have been used to argue that climate models, which do not specifically account for the energy of mixing, potentially miss a very important feedback on climate change. In this paper, we re-examine the question of what energetic arguments can teach us about the climate system and conclude that the relationship between energetics and climate is not straightforward. By analysing the buoyancy transport equation we are able to demonstrate that the large-scale transport of heat within the ocean requires an energy source of around 0.2 TW to accomplish vertical transport and around 0.4TW (resulting from cabelling) to accomplish horizontal transport. Within two general circulation models this energy is almost entirely supplied by surface winds. We also show that there is no necessary relationship between heat transport and mechanical energy supply.

1 Introduction

This paper examines the linkage between mechanical energy supply and thermal energy transport associated with the ocean circulation. The large-scale ocean circulation plays an important role in maintaining the earth's climate. Recent estimates of heat transport show that the oceans export 3.2 petawatts from the tropics, as shown by the stars in Figure 1a (Trenberth and Caron, 2002). In the absence of this heat flux, the high latitudes would cool significantly. Indeed, recent work suggests that without this flux of heat the entire world would freeze over as sea ice spread equatorwards (Winton, 2003).

In a seminal paper, Munk and Wunsch (henceforth MW98) argued that one could use the mechanical energy budget to draw conclusions about what mechanisms were responsible for driving this circulation. The abstract of

MW98 concludes with the statement "a surprising conclusion is that the equator-to-pole heat flux of 2000 TW associated with the meridional overturning circulation would not exist without the comparatively minute mixing sources. Coupled with the findings that mixing occurs at a few dominant sites, there is a host of questions concerning the maintenance of the present climate state, but also that of paleoclimates and their relation to detailed continental configurations, the history of the Earth-Moon system, and a possible great sensitivity to details of the wind system." MW98 pose a number of intriguing questions, including whether tidal mixing puts a lower limit on heat transport, or whether it is constrained by air-sea fluxes.

The results of MW98 have spurred much interest in the role of internal tides in producing intense mixing, leading to extensive field programs such as the Hawaii Ocean Mixing Experiment (Rudnick et al., 2003), as well as recent modeling studies of ocean tide generation and energy conversion (Simmons et al., 2004; Arbic et al., 2004a). This work is yielding a great deal of insight into how turbulent mixing within the ocean is generated, and is clearly important for understanding the deep ocean circulation.

However, the necessity for some sort of mechanical mixing to drive the ocean overturning has also spurred a number of authors to consider the mechanical energy flux as a sort of "control knob" on the global overturning. Huang (1999, henceforth H99) showed that in idealized models of the meridional overturning the dependence of heat transport and meridional overturning on temperature gradient differed between models which kept dissipation constant and models that kept the diffusion coefficient constant. Emanuel (2002) suggested that the input of mechanical energy by tropical cyclones could represent an important stabilizing feedback on climate. Wunsch (2003) suggested that tidal amplitudes during the last glacial maximum were higher than at present (a suggestion supported by recent modeling studies by Egbert et al. 2004 and Arbic et al, 2004b) and argued that such higher tidal amplitudes should have led to an enhanced meridional overturning circulation-in contrast with the standard picture of weaker overturning during this period.

In this paper, we argue that analyzing the ocean circulation in terms of mechanical energy supply leads to incorrect intuitions about the sensitivities of the circulation. In particular, we show that

1. Only a small fraction of the 2TW estimated by MW98 to drive the whole ocean overturning is in fact required to explain the *lateral heat transport*.
2. This fraction is most likely supplied by the winds.
3. Increasing the supply of mechanical energy does not necessarily imply an increase in the heat transport.

We do this by looking in detail at the buoyancy transport equation. In Section 2, we use this equation to clarify what it means to say that the ocean is not a convective system, to identify the key processes that move density in the vertical, and to estimate which of these processes are really important in producing the observed lateral heat transport. In Section 3 we describe two general circulation models that produce reasonable distributions of tracers. In Section 4 we examine the buoyancy transport within these models.

2 What the buoyancy transport equation tells us about ocean energetics

2.1 Buoyancy flux and the energetics of the circulation

The fact that a key facet of the the ocean general circulation is the sinking of cold, dense waters in high latitudes led a number of investigators over the years to consider it a form of convection (Stommel, 1961; Huang et al., 1992; Park and Whitehead, 1999). In this view, the ocean is analogous to a pot of water simmering on a stove in which hot water rises along the edges, is cooled as it moves inwards at the surface, and sinks in the middle. In the Boussinesq approximation of such a circulation, parcels gain buoyancy $b = -g\rho/\rho_0$ (where g is the gravitational acceleration, ρ the in-situ density and ρ_0 is a mean density) at the bottom of the pot, and lose it at the top. Let w represent the vertical velocity and let $<>$ denote a horizontal integral and $\hat{}$ a temporal average, In a convective system the integral

$$\frac{1}{T}\int\int\int\int \rho_0 \times w \times b \; dxdydzdt = \rho_0 \int < \widehat{wb} > dz = P_{buoy} > 0 \tag{1}$$

since updrafts ($w > 0$) carry positive buoyancy and downdrafts ($w < 0$) carry negative buoyancy. P_{buoy} represents the power released by convection or the rate of buoyancy work. Buoyancy work drives the mean circulation in a simmering pot, the earth's mantle, and in the atmosphere. In a convective system with constant viscosity μ in which there is no work done by the boundaries or internal sinks and sources of buoyancy

$$\rho_0 \int < \widehat{wb} > dz = \rho_0 \int < |\mu \widehat{\nabla \mathbf{u}}|^2 > dz \tag{2}$$

This means that an increase in the left side of the equation implies larger dissipation and a faster circulation. The mechanical energy budget of a convective system is thus a useful measure of the circulation.

For many years, however, an argument has raged about whether the ocean circulation can in fact be maintained by surface buoyancy forcing. Sandstrom (1908) came to a conclusion which is restated in H99 as follows: "a closed steady circulation can be maintained in the ocean only if the heating source is situated above the cooling source." As discussed by H99 the applicability of this theorem to ocean circulation has been debated over many decades. One problem is that Sandstrom's theorem is derived for a model system that differs in significant ways from the real ocean, ignoring diffusion, friction, and salinity. Given that Park and Whitehead (1999) present a laboratory model of the thermocline which reproduces many features of modeled overturning circulations but which violates Sandstrom's theorem (heating and cooling being situated at the same level) it is far from clear that the theorem should give any insight into the real ocean.

In what follows, we point out that the buoyancy transport equation offers a simpler way of thinking about the energetics of the large-scale circulation.

We begin by considering the density transport equation

$$\frac{\partial \rho}{\partial t} + \frac{\partial}{\partial x}(u\rho) + \frac{\partial}{\partial y}(v\rho) + \frac{\partial}{\partial z}(w\rho) = Q_\rho \tag{3}$$

where Q_ρ is a source term for density. For our purposes, we will suppose that it includes internal sources due to nonlinearities of the equation of state as well as viscous heating and molecular diffusion. All other turbulent fluxes are assumed to be contained in the advection terms. If we integrate this from the bottom of the ocean to some horizontal surface the horizontal transport terms drop out and take a temporal average (we are interested in the steady-state flow) we obtain

$$< \widehat{w\rho} >= \int_{z=-D}^{z_0} < \widehat{Q_\rho} > dz \tag{4}$$

Multiplying by $-g/\rho_0$, we obtain

$$< \widehat{wb} >= \int_{z=-D}^{z_0} -g < \widehat{Q_\rho} > /\rho_0 dz = \int_{z=-D}^{z_0} < \widehat{Q_b} > dz \tag{5}$$

where Q_b includes such terms as geothermal heating and cabelling (Huang, 2004). In what follows, we will use both numerical models and data to estimate some of the terms that go into making up both the left and right-hand sides of this equation.

2.2 Analyses which assume small interior buoyancy sources and sinks

We begin by looking at approximate solutions of this equation that hold when Q_b is taken to be negligible. We note that this assumption is fundamental to previously published results such as MW98 and H99. Then

$$< \widehat{wb} > \approx 0 \tag{6}$$

What does this mean? First, it means that the ocean is not a convective system *in the sense that buoyancy work does not provide energy to maintain the flow*. MW98 and H99 also argue that the ocean is not a convective system but do not link it directly to buoyancy transport. If vertical buoyancy transport is zero, then buoyancy work is also zero. While this does not mean that the actual flow is zero (Papparella and Young, 2002), it does imply that there must be compensation between transports associated with different spatial and temporal scales. If the large-scale, time-mean flow brings buoyancy upwards, some other scales must act to move it downwards. In order for such flows to exist in the presence of dissipation, however, there must be some external source of energy to the system. In the laboratory experiment of Park and Whitehead (1999) it is the internal energy of the system which is released by molecular diffusion. In the real ocean, the most important external source of energy is mechanical work from winds and tides.

This does not mean that buoyancy is unimportant in the system. Changes in the surface buoyancy distribution resulting from changes in the net fresh-water balance can alter the geometry and magnitude of the circulation (Bryan, 1986; Gnanadesikan, 1999; Seidov and Haupt, 2003; Saenko et al., 2003; Saenko and Weaver, 2004). However, as long as the buoyancy flux is *not positive* buoyancy work is not the important source of energy for oceanic circulation that it is for atmospheric circulation.

2.3 Decomposing the buoyancy transport equation

One can gain more insight into the energetics of the system by decomposing the buoyancy transport into four terms roughly corresponding to the time scales involved in the vertical velocities.

$$< \widehat{wb} > \equiv < \widehat{w}\widehat{b} > + < \widehat{w_e b_e} > + < \widehat{w_c b_c} > + < \widehat{w_t b_t} > = \int_{z=-D}^{z_0} < \widehat{Q_b} > dz$$

$$\equiv \text{ Mean flow+Mesoscale eddies+Convection+Small-scale Turbulence= Buoyancy sinks} \quad (7)$$

Essentially, the first term is the long-term, large-scale mean, the second is associated with spatial scales of tens of km and temporal scales of days, the third with spatial scales of tens to hundreds of meters and temporal scales of minutes, and the fourth with spatial scales of cm and temporal scales of seconds. In the event that $Q_b = 0$ (something that we show is not in fact the case in Section 4) any of these terms can be nonzero, only their sum must vanish. Integrating these terms over the volume of the ocean yields the total work associated with each of them.

In coarse-resolution general circulation models used in climate studies (Griffies et al., 2001) these terms are usually represented by separate routines. The first is handled by the tracer advection routines, The second is handled by isoneutral diffusion schemes (Gent and McWilliams, 1990; Griffies et al., 1998; Griffies, 1998). The third term is dealt with by convective adjustment and the fourth by parameterizations of small-scale vertical diffusion in terms of some mixing coefficient (Bryan and Lewis, 1979; Gnanadesikan et al.,2002).

This decomposition can be justified if one thinks of the various processes as occupying different locations in wavenumber-frequency space. Insofar as we are looking at the long-term mean and globally integrated budgets, only those components which have nearly identical frequencies and wavenumbers will contribute to a spatiotemporal average. This is particularly important insofar as we are considering coarse models, which essentially assume some separation between advection on scales of the grid and that on scales of the much smaller mesoscale eddies. Our decomposition would be much more difficult if we were considering models which resolved the advective flows associated with mesoscale eddies.

2.4 Energetic consequences of this decomposition

The argument of MW98 can be recovered from (7) by setting the mesoscale eddy, convective terms, and buoyancy sink terms to zero. They then assume a circulation scheme in which dense water sinks into the deep ocean, becomes light as a result of downward buoyancy flux associated with small-scale turbulence and upwells at a lighter density. Equation 8 then becomes

$$F_{buoy} = < \widehat{w}\widehat{b} > = - < \widehat{w_t b_t} > = K_v < \widehat{N^2} > = \gamma < \widehat{\epsilon} > /\rho_0 \tag{8}$$

where K_v is a turbulent diffusion coefficient, N is the buoyancy frequency, $\gamma \approx 0.2$ (Oakey et al., 1982; Polzin et al., 1995) is a turbulent efficiency, and ϵ is the turbulent kinetic energy dissipation in W/m^3. Given 30 Sv of water injected to a depth of 4km and rising to a depth of 1 km with a density difference of 1 kg m^{-3} implies that the buoyancy flux profile F_{buoy} goes from $3 \times 10^5 m^4/s^3$ at 1 km to 0 at 4 km. The energy required to produce such a flux profile (given the low efficiency of turbulent mixing) is $\rho_0 * F_{buoy} * 3000m/2 * 5 = 2.25$ TW. This number is so much larger than the 0.9 TW supplied by the winds working against the geostrophic current (Wunsch, 1998) that it implies a significant source of mechanical energy is needed to supply ϵ. MW98 use this discrepancy to argue that tides could affect climate. Webb and Suginohara (2001) have criticized this argument on the grounds that much of the water injected into the deep ocean does not cross isopycnals but is upwelled in other parts of the ocean. We will return to their argument later in this paper.

In the meantime it is worth asking whether the decomposition of MW98 is the right one for looking at the circulation that actually accomplishes the transport of heat within the ocean. In what follows, we argue that a different balance is involved, invoking the following train of reasoning.

1. Heat transport must involve the loss of heat in high latitudes.

2. This heat loss is associated with convection.

3. Convection extracts heat from (on average) the middle of the mixed layer and brings it to the surface. This is associated with an upward flux of buoyancy.

4. In order for $<\widehat{wb}> \approx 0$ there must be some compensating downward

fluxes of buoyancy, either from turbulence or large-scale advection at other locations or times.

5. These compensating downward fluxes of buoyancy require energy.

6. Thus by estimating the upward flux of buoyancy associated with convection, we can estimate the energy required to balance this buoyancy flux, and thus to drive the ocean heat transport.

The upward buoyancy flux associated with the convective term in (7) can be calculated as follows. By definition, within a mixed layer the temperature, salinity, and buoyancy are well mixed and change coherently. Since in one dimension, $\partial b/\partial t = -\partial/\partial z(\overline{wb})$, assuming we have a mixed layer is identical to assuming that the vertical flux divergence is constant in z. Thus within a mixed layer (if one takes an appropriate *local* spatial average - over many individual convective cells), one can approximate $\overline{wb}$ as varying linearly between the surface buoyancy flux and zero at the mixed layer base D_{ML}. Then $\overline{wb} = \overline{wb}|_{z=0}(z + D_{ML})/D_{ML}$. The surface buoyancy flux associated with a surface heat flux Q is just $g\alpha Q/c_p$ where $\alpha = (1/\rho)\partial\rho/\partial T$ is the coefficient of thermal expansion and c_p is the specific heat. Integrating $\overline{wb}$ over the mixed layer then gives us equation (9).

$$\int \rho\overline{wb}dz = \frac{g\alpha D_{ML}}{2c_p}Q = Q_{\text{mech}} \tag{9}$$

where Q_{mech} has units of Wm^{-2} and represents a mechanical energy flux. A standard interpretation of Q_{mech} (see for example Gill and Turner, 1976) is that it is the energy flux required to stir the mixed layer (when $Q > 0$) or released by convection (when $Q < 0$). However, in the context of equation (7) it represents a nonzero *local* contribution to a *global* buoyancy budget which must be balanced by a buoyancy flux of the opposite sign somewhere else within the domain (or potentially at the same spot, but at a different time). We can thus define $Q_{\mathrm{con}} = Q_{\mathrm{mech}}$ $(Q < 0)$ as a *convective energy demand* associated with the heat transport. The globally integrated convective energy demand is then

$$- < \widehat{Q}_{\mathrm{con}} >= \int \rho_0 (< \widehat{w}\widehat{b} > + < \widehat{w_e b_e} > + < \widehat{w_t b_t} >) dz \tag{10}$$

We refer to this as an energy demand because it represents a constraint on the energy flux that must be put into the system to drive the flows on the right-hand side of (11). These flows move buoyancy downwards so that it can be moved back up again where convection is occurring.

Similarly, we can define the *mixed layer potential energy demand* $Q_{\mathrm{mix}} = -Q_{\mathrm{mech}}$ $(Q > 0)$ as the energy flux needed to stir heat down into the mixed layer in regions where the ocean is gaining heat. Q_{mix} is that part of $/int < w_t b_t > dz$ which is due to wind-driven deepening of stable mixed layers. Note that when D_{ML} is small (say 100m), $|Q_{\mathrm{mech}}|/|Q| \sim \times 10^{-5}$! That is, the mechanical energy flux associated with moving heat from point to point in the ocean is very much smaller than the thermal energy flux involved.

We can obtain an estimate of the convective energy demand and mixed layer potential energy demand by examining observed heat fluxes and mixed layer depths. Figure 2a shows the Q_{con} when surface fluxes Q are given by the dataset of Josey (1999) and mixed layer depths D_{ml} given by the World Ocean Atlas (Levitus and Boyer, 1998). A Q_{con} value of 10 mW/m^2 does not imply that there must be a local flux of energy to the ocean. Figure 2a simply shows what individual regions contribute to the globally integrated demand. The global integral of these energy demands is shown in Table 1, and compared with the fluxes associated with wind stress.

The globally integrated convective energy demand is only about 0.15 TW, while the globally integrated mixed layer potential energy demand is about 0.2 TW. This value is almost an order of magnitude *smaller* than the work done by the winds on the general circulation. In the absence of cabelling, the convective energy demand is a direct estimate of the amount of energy needed to drive the true circulation. The key point we would make here is that these numbers do not seem to *require* a source of tidal mixing.

The difference between our estimate of 0.15 TW and the MW98 estimate of 2TW arises from our focus on the work associated with the heat transport — which is largely confined to the surface ocean. Recent work by Talley (2003) argues that the deep circulation transports 0.14 PW southwards across 30S, with southward transports decreasing as we move northwards. This is only 5% of the tropical heat export. Since MW98 consider the buoyancy work only within the deep ocean, they effectively ignore those flows that do most of the heat transport. From equation (8), the energy flux required to drive the deep circulation Q_{deep} is

$$Q_{deep} \equiv \int < \epsilon > dz \approx 5 * \rho_0 * < \widehat{w}\widehat{b} > |_{z=1km} * 1500m * A_1 \qquad (11)$$

where A_1 is the area of the ocean at a depth of 1km. The 2 TW associated with the deep circulation comes about because volume of integration is much larger than for Q_{con} and Q_{mix}, and because the efficiency γ is low. This illustrates why thinking about the ocean circulation in terms of energy is not straightforward. Not all energy inputs have equivalent impacts on heat transport.

It might be argued that our result is not significantly different from MW98 in that the low mixing efficiency within the mixed layer demands a large (of order 1 TW) energy flux to satisfy the mixed layer potential energy demand. However, the key point of MW98 is their argument that the mechanical energy be supplied at great depths, away from the ocean surface- *requiring an input of tidal energy*. In constrast to the ocean interior, the mixed layer has access to substantial sources of turbulent kinetic energy. For example, the direct turbulent input to waves has been estimated as 100 ρu_*^3 (Agrawal et al., 1992) where $u_* = (\tau/\rho)^{1/2}$ is around 0.01 to 0.02 m/s for most of the oceans. A rough estimate we made using winds taken from the NCEP reanalysis data set (Kalnay et al., 1997) showed that this term was of order 100 TW, much larger than needed to mix the surface ocean. Wang and Huang (2004a,b) estimate the flux of energy to inertial oscillations at 3 TW and the flux to surface waves at 60 TW. Thus, in contrast to MW98, our budget thus far does not imply that there is "missing mixing" that must be supplied by the tides.

2.5 Cabelling and the buoyancy equation

The argument made up to the present point has one major flaw-namely that it ignores the role of cabelling. When two water parcels of equal mass are mixed, the resulting water is denser than either one. This means that sources of buoyancy can be associated with the *lateral transport* of buoyancy so that if $F^{x,y}_{buoy,T,S}$ are the transports of buoyancy, temperature, and salinity in the x and y directions respectively

$$< \widehat{w}b > |_{z=z_r} = \int_{-D}^{z_r} - < \frac{\partial}{\partial x}\widehat{F^x_{buoy}} - \frac{\partial}{\partial y}\widehat{F^y_{buoy}} - \frac{g\widehat{Q_\rho}}{\rho_0} >=$$

$$-g\int_{z=-D}^{z_r} < \alpha\frac{\widehat{\partial}}{\partial x}F^x_T + \beta\frac{\widehat{\partial}}{\partial x}F^x_S > ++ < \alpha\frac{\widehat{\partial}}{\partial y}F^y_T + \beta\frac{\widehat{\partial}}{\partial y}F^y_S > \qquad (12)$$

assuming no internal sources of heat and salinity. Insofar as the horizontal circulation is picking up heat in areas where the temperature is high (and thus α is large) and losing it in areas where temperature is low (and α is small), the lateral transport of buoyancy can result in a nonzero buoyancy source. In order to get a better estimate of these terms, we now turn to two numerical general circulation models in which all the terms we have discussed so far can be calculated explicitly.

3 Model description

The models used in this paper are implemented using the Geophysical Fluid Dynamics Laboratory Modular Ocean Model, v.3. (Pacanowski and Griffies, 1999). The model is run at a nominal resolution of 4.5° in latitude and 3.75° in longitude with 24 staggered vertical levels ranging from 25m thick at the surface to 450m thick at depth. Two implementations of the model were run:

1. PRINCE2: Model PRINCE2 is built from the KVHISOUTH+AILOW described in Gnanadesikan et al. (2002). In model KVHISOUTH+AILOW, the base topography (adopted from earlier versions of the GFDL coupled climate model) has a wide Drake Passage. Wind stresses are given by the dataset of Hellermann and Rosenstein (1983). Surface heat and salt fluxes are derived by a combination of applying heat fluxes from the data set of daSilva et al. (1994) and restoring the surface temperature and salinity to the monthly Levitus ocean atlas (1994) with a time scale of 30 days. Vertical diffusion is given by the profile of Bryan and Lewis (1979), going from 0.15 cm^2/s in the surface ocean to 1.3 cm$^2/s$ in the deep ocean with a relatively large value (1.0 cm ^{2}/s) at all depths in the Southern Ocean (Polzin, 1999).

   Model PRINCE2 makes the following changes to KVHISOUTH+AILOW. First, at four grid points around Antarctica during the winter months,the restoring salinities are changed so as to ensure that the observed values of Weddell and Ross Sea bottom waters are actually found at the surface. Secondly, the value of the vertical diffusivity within the pycnocline is increased from 0.15 to 0.3 cm^2/s.

2. PRINCE2A: Model PRINCE2A has the same changes to the surface restoring around Antarctica as model PRINCE2, but it does not include the change in vertical diffusivity. Instead, the surface wind stresses are changed from the Hellermann wind stress product to the ECMWF analysis (Trenberth et al, 1989), which gives higher wind stresses in the Southern Ocean and lower wind stresses in the tropics. Drake Passage is narrowed by one grid box to make it more realistic.The vertical diffusivity is increased in the top level of the model to produce more realistically deep mixed layers. Finally, it was found that the "flux corrections" computed by restoring surface salinity and temperature to observations in many locations were in the opposite direction of the applied fluxes. This was particularly true in the Southern Ocean. The (apparently biased) applied fluxes were changed by adding the restoring correction computed from a 400 year-long run.

Figure 3 shows that the models reproduce the horizontally-averaged temperature, salinity and radiocarbon distributions quite well. The errors in all three fields are small in comparison to the observed range and all major features are captured. Both models represent credible solutions for the ocean circulation, with reasonable rates of vertical exchange (more analysis of the vertical exchange in these models is presented in Gnanadesikan et al., 2004).

Although there are some small biases with respect to the salinity, these have a minor impact on the overall stratification, which is of the most importance when energetics are considered and is very close to observations.

4 General Circulation Model Results

In addition to the observational estimate of lateral heat transport, Figure 1 shows heat transports in PRINCE2 and PRINCE2A. Heat transport in the PRINCE2 model is very close to the observations (Table 1), while the PRINCE2A model has a much weaker lateral heat transport. It might be expected that the increased diffusivity in PRINCE2 was the primary driver of the 1.1 PW difference in tropical heat export. However, examination of the heat transport in the model suite from which these runs were spun off shows that even with low values of diffusion, the Hellermann winds produce a heat export of 2.95 PW. Moreover, a much larger increase in vertical diffusion to 0.6 $\mathrm{cm^2 s^{-1}}$ produces an increase of heat export of 0.9 PW, leading us to conclude that the change in diffusion accounts for at most 1/4 of the difference between PRINCE2 and PRINCE2A (Gnanadesikan et al., 2003). Zonal averages of Q_{con} and Q_{mix} from the models are compared with the observational estimates in Figure 4 and show good qualitative agreement.

One advantage of the models is that the detailed vertical budget of heat and buoyancy can be calculated from all the terms. Figure 5 shows the implied buoyancy flux computed from integrating the terms that comprise the right-hand side of equation (13). A somewhat startling result is that the buoyancy flux is *not* 0 but is in fact large and negative. Far from being driven by buoyancy, the ocean circulation actually results in a buoyancy sink! Integrating the implied flux gives a value of 0.44 TW for model PRINCE2 and 0.42 TW for model PRINCE2A. The internal loss of energy due to the nonlinearity of the equation of state is actually a major component of the buoyancy budget.

What vertical fluxes balance this term? Does the introduction of this source require the reintroduction of a strong downward diffusive flux of buoyancy? In order to answer this question we examine the vertical fluxes of heat (Figure 6a), salt (Figure 6b) and the buoyancy fluxes associated with heat and salt (Figures 6c and 6d), decomposing them into components due to advection, convection, and subgridscale diffusion. A number of important results emerge from this decomposition.

The first is that the total vertical advective flux of heat (Figure 6a) and buoyancy (Figure 6c) (corresponding to $< \widehat{w}\widehat{b} >$ in (7)) is *downwards* in both the models (the temperature term dominates the buoyancy flux). It is this downward advection of heat, not the upward flux due to subgridscale parameterization of mixing (corresponding to the sum of the eddy and turbulence terms in 1) that primarily balances convection and cabelling. A similar result was noted for heat fluxes in one previous model study (Gregory, 2000) but the implications for ocean energetics were not explored.

The fact that the advective heat transport is downward (and that as a result so is the buoyancy transport) has important implications. If the ocean is hydrostatic and velocities vanish on the boundaries (as is the case in these models), then if p is pressure the buoyancy work must be equivalent to the work done by the horizontal pressure gradients on the mean flow.

$$\int \rho < \widehat{w}\widehat{b} > dz = \int < \widehat{w}\frac{\widehat{\partial p}}{\partial z} > dz = -\int < \frac{\widehat{\partial w}}{\partial z}\widehat{p} > dz =$$
$$\int < \widehat{\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}}\widehat{p} > dz = -\int < \widehat{u}\frac{\widehat{\partial p}}{\partial x} + \widehat{v}\frac{\widehat{\partial p}}{\partial y} > dz \qquad (13)$$

A downward transport of heat and buoyancy implies that this must be negative, so that pressure gradients must work *against* the mean flow. Geostrophic flow is by definition along the pressure gradient, and so does not contribute to the pressure work. Frictional flows driven by pressure gradients move from high to low pressure, resulting in positive rather than negative pressure work. Only the wind-driven flow in the mixed layer, which converges water into the subtropical highs, represents a large source of negative pressure work. Ekman pumping is the dominant driver of the buoyancy budget in realistic GCMs.

Note that if we integrate over the mixed layer, where we assume pressure gradients and Ekman flows are essentially constant, we get

$$< (\tau^y p_x/\rho f) - \tau^x p_y/\rho f >=< \vec{\tau} * \vec{u}_g > \qquad (14)$$

so that the pressure work is equivalent to the work done by the winds on the geostrophic current (Wunsch, 1998). In our models surface winds are not only sufficiently energetic to drive the heat transport, *they are the only process that has the correct sign* to explain the advective fluxes of heat and buoyancy and to balance convection and cabelling.

A second important result is that the mechanical energy supply is not a good predictor of the lateral heat transport. In fact, the model with the larger vertical transport of heat (and thus the large convective energy demand) has the *smaller* lateral transport of heat. Additionally, despite the large difference in the heat transports, there is relatively little difference in the cabelling sink of energy. As can be seen in Figure 7 (which shows $\tau * u$) in the two models, PRINCE2 has a larger input of mechanical energy in the tropics, while PRINCE2A is larger in the Southern Ocean. The Southern Ocean dominates the global input of wind energy so that PRINCE2A exhibits a larger vertical transport of heat (as would be expected from previous work suggesting that the large scale overturning circulation is wind-driven). However, the small change in energy input in the tropics is more important for lateral heat transport. This is because the lateral flows in the mixed layer associated with wind stress scale as $\tau/sin(latitude)$, and so small changes in absolute magnitude of the wind stress at the equator have a much larger impact than large changes in the Southern Ocean. Once again, not all energy inputs are equal.

A third important result is that the subgridscale fluxes are essentially equal in both models, despite the fact that one model has a higher vertical diffusion than the other. This result points out the important role played by parameterized mesoscale eddies in these models. Mesoscale eddies act to flatten isopycnal surfaces (Gent and McWilliams, 1990; Gnanadesikan, 1999). This results in an "eddy-induced" advective flow in which cold, dense water descends and warm, light water ascends. In the thermocline the dominant balance is between downward advection of heat associated with the mean flow, and upward advection of heat induced by the eddies. This can be seen in Figure 8, which presents a breakdown of the heat budget into subgridscale diffusive and advective terms. The eddy-induced advective terms constitute the dominant subgridscale mixing terms above 1500m and essentially compensate the diffusive fluxes below that depth.

A final important result is that the horizontally averaged advective flux of heat and buoyancy is actually negative down to 2500m in the models. The classic picture of upward advection and downward diffusion of heat (Munk, 1966) holds only over a fraction of the ocean and is associated with relatively weak fluxes of heat. A similar point is made by Gregory (2000) who finds the classical balance to hold in the tropical thermocline, but not globally. The MW98 estimate that 2TW of energy are needed to maintain the deep stratification against upwelling assumes that a large amount of water (30 Sv) upwells through the deep stratification.However it is not clear that such large fluxes are actually necessary. Webb and Suginohara (2001) make the point that the actual buoyancy flux in the deep ocean may be quite weak, with only 5-6 Sv of deep water upwelling across isopycnals rather than the 30 Sv of MW98. Vallis (2000) finds that an idealized model of ocean circulation can support deep stratification as abyssal mixing goes to zero, as the flow across the stratification also goes to zero. Insofar as these pictures actually describe the deep ocean, tides need not do a lot of work in the deep ocean.

This implies that winds, rather than unresolved tidal processes, play the dominant role in climate.

5 Caveats

In order to keep the argument relatively straightforward we have neglected certain aspects of the solution. In this section, we consider two points that we have neglected.

The first point is the neglect of geothermal heating. How important is this neglect? Estimates of the rate of geothermal heating are around 50 mW/m^2 (Huang, 2004) or a heat flux of 18TW. If we assume this to be injected at 4000m depth (an overestimate as significant amounts of injection occur along ridge crests), the associated energy flux is $g\alpha/c_p * 4000\text{m} * 18\text{TW}$ or 0.036 TW- a significant number in comparison with the small energy fluxes in the deep ocean but a small contributor to the overall mechanical energy budget.

A second caveat relates to our use of a linear buoyancy flux profile within the mixed layer. In fact, large eddy simulations (Large et al., 1994) have been used to argue that the buoyancy flux profile actually has the form

$$\overline{wb} = \overline{wb}(z=0) * (1.2 * z + D)/D \quad z > -D \tag{15a}$$

$$\overline{wb} = -\overline{wb}(z=0) * (z + 1.2D)/D \quad z > -1.2D \tag{15b}$$

so that the turbulence at the surface entrains lighter water from below the mixed layer. This reduces the convective energy demand from $0.5g\alpha QD_{ml}/c_p$ to $0.42g\alpha QD_{ml}/c_p$. The turbulent convective energy generated by convection is able to mix some buoyancy downwards-partially satisfying the convective energy demand.

6 Conclusions

We have presented a decomposition of the vertical buoyancy equation that casts a new light on the relationship between the mechanical energy supply to the ocean and the lateral transport of thermal energy by the ocean. The approximately 3 PW of lateral heat transport involves relatively small vertical excursions. As a result it requires a very small input of mechanical energy to move buoyancy downwards so as to balance convection (0.15-0.2 TW) and a surprisingly large input of mechanical energy to balance cabelling (0.4 TW). In comprehensive general circulation models this energy is efficiently supplied by the winds. This has important implications for climate models. Insofar as coupled models of climate change capture changes in wind stress, they would be expected to capture changes in heat transport as well. The deep circulation, by contrast, involves very small fluxes of heat which are inefficiently driven by diffusion over large vertical scales. While such circulation may require a significant flux of energy (and thus be strongly influenced by tidal mixing), it does not directly affect the lateral transport of heat. Thus, while mechanical energy supply is an interesting diagnostic of the ocean circulation, it depends on the vertical scale of circulation as well as the heat flux carried by the circulation. This means that it cannot be used to predict lateral heat transport and its impact on climate.

It is possible that deep mixing may play a role in climate, but this role is likely to be indirect. In the Southern Ocean, upwelling of warmer Circumpolar Deep Waters plays a role in determining sea ice extent and thus surface albedo. The rate of vertical exchange in the Southern Ocean also has important implications for determining atmospheric carbon dioxide (Marinov, 2005; Toggweiler and Russell, manuscript in prep.). Insofar as deep mixing can affect the properties of the deep ocean, it may play a role in such climatically important processes- it is impossible to draw more robust conclusions without more evidence. However, it is clear that the location and magnitude of the mean wind stress excercises a primary control on the oceanic heat transport, and thus on global climate.

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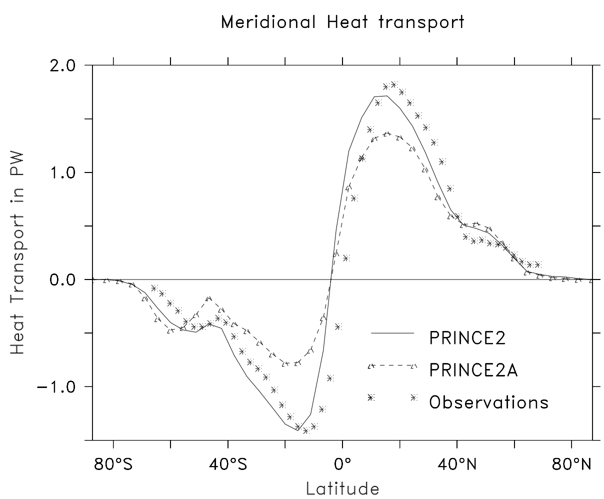

Figure 1: Lateral and vertical heat transports. North-south heat transport in data (stars) and two coarse-resolution general circulation models.

	Data	PRINCE2	PRINCE2A
Tropical Heat Export (PW)	3.23	**3.12**	**2.15**
Convective Work Demand (TW) (Estimated from heat flux)	**0.15**	**0.19**	**0.28**
Mixed Layer Work Demand (TW) (Estimated from heat flux)	**0.2**	**0.07**	**0.14**
Direct Wind Input ($\vec{\tau} * \vec{u}_g$, TW)	0.77,0.88	**0.66**	**0.85**
Total wind input ($\vec{\tau} * \vec{u}$)		**1.03**	**1.18**
Direct Wind Input (South of 20S)	0.63	**0.51**	**0.64**
Direct Wind Input (20S-20N)	0.13	**0.08**	**0.12**
Advective work (TW)		**0.49**	**0.69**
Subgridscale work (TW)		**0.09**	**0.03**
Cabelling demand (TW)		**0.44**	**0.42**
Convective demand (full, GCM)		**0.15**	**0.20**

Table 1: Energy transports, sources and sinks in data and models. Data reported here for the first time are shown in boldface. "Observed" tropical heat exports are taken from Trenberth and Caron (2001).The higher of the observed global direct wind input numbers is taken from Wunsch (1998) as are the regional numbers. The lower value is taken from Scott (1998). The convective work demand and mixed layer work demand are the global integrals of the convective energy demand and mixed layer energy demand based on temperature alone. The convective and cabelling sinks are computed directly from the appropriate terms within the models. Note that while PRINCE2 balances well (sum of demand terms is approximately equal to sum of work terms), PRINCE2A does not balance exactly.

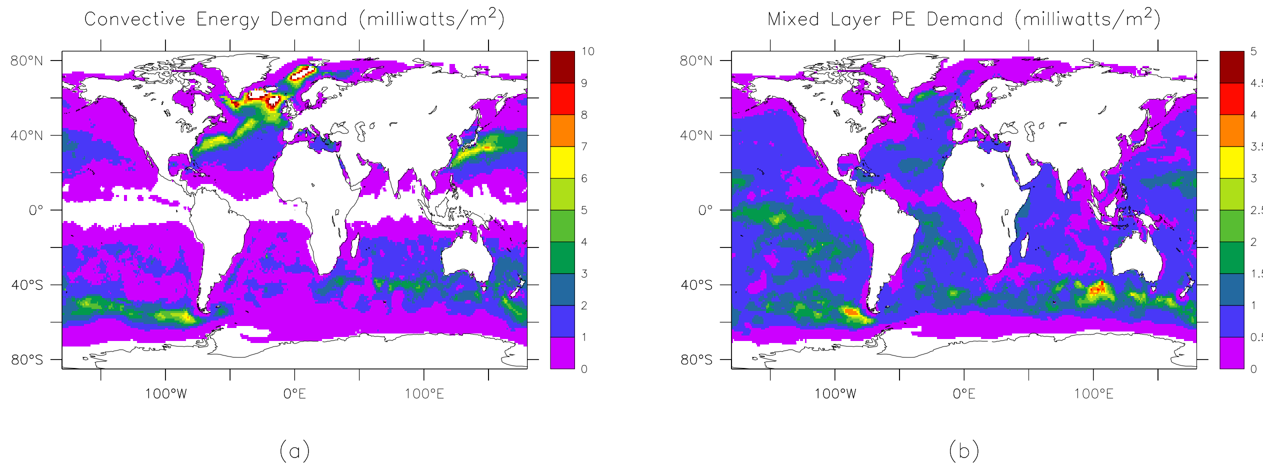

Figure 2: Energy fluxes in millwatts/m^2 implied by surface fluxes and mixed layer depths. (a) Mechanical stirring required to supply convection when there is net cooling. (b) Mechanical stirring needed in regions with net heating to homogenize mixed layer depth.

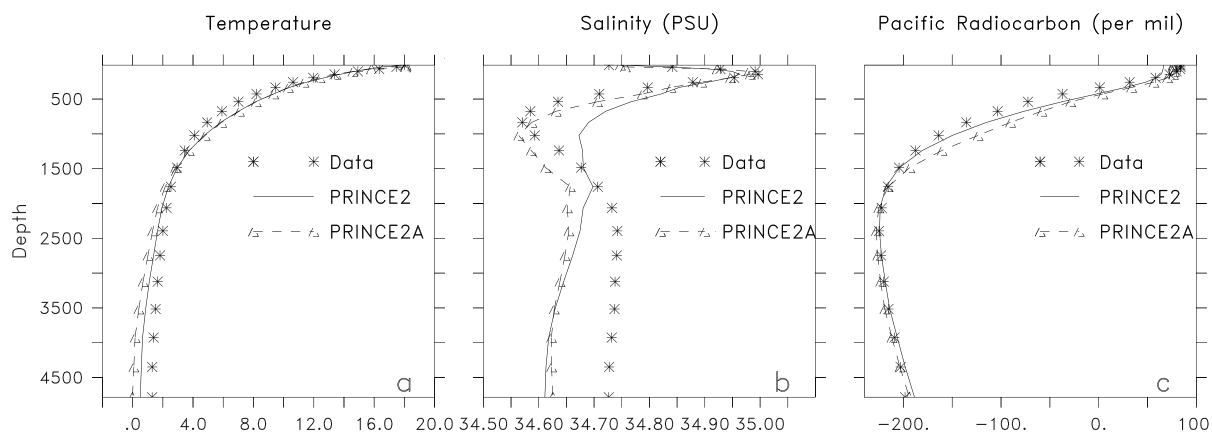

Figure 3: Demonstration that the two models presented here have a reasonable representation of the large-scale ocean structure. (a) Horizontally averaged temperature in data (stars) and the models. (b) Horizontally averaged salinity. (c) Radiocarbon in per mil averaged over the Pacific Sector (110E to 100W and 60S to 60N).

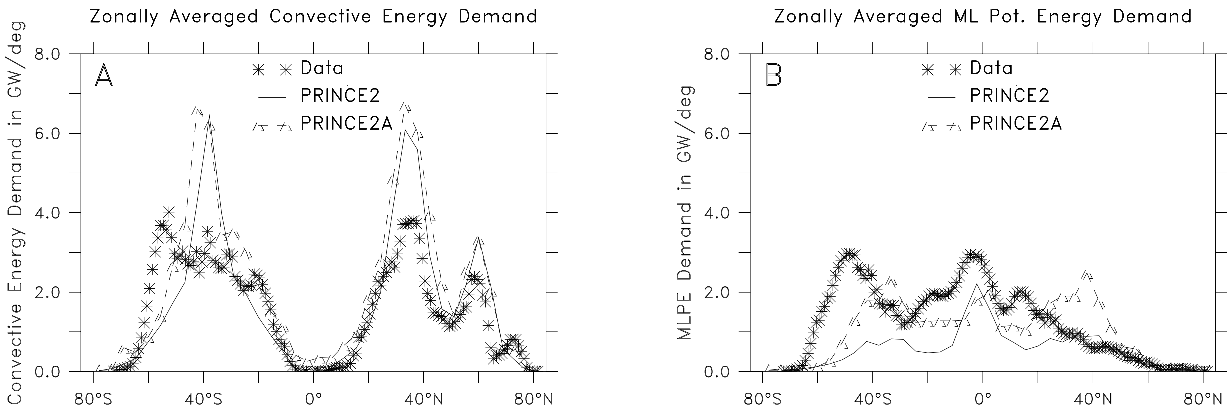

Figure 4: (a) Zonally integrated convective energy demand in data and the two models. (b) Zonally integrated mixed layer PE demand in the two models.

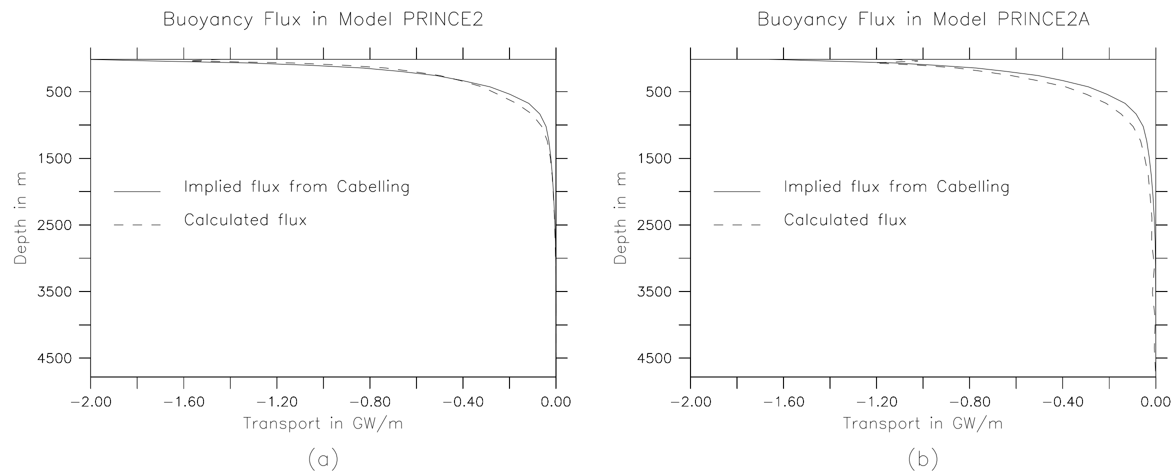

Figure 5: Vertical buoyancy balance in the models between cabelling terms and vertical transport. Exact agreement is not expected both because of cabelling due to vertical mixing terms and inaccuracies due to numerical truncation.(a) Vertical buoyancy flux in model PRINCE2 (solid) and implied flux from integrating the horizontal mixing and advection terms up to the same depth (dashed line). (b) Same as (a) but for PRINCE2.

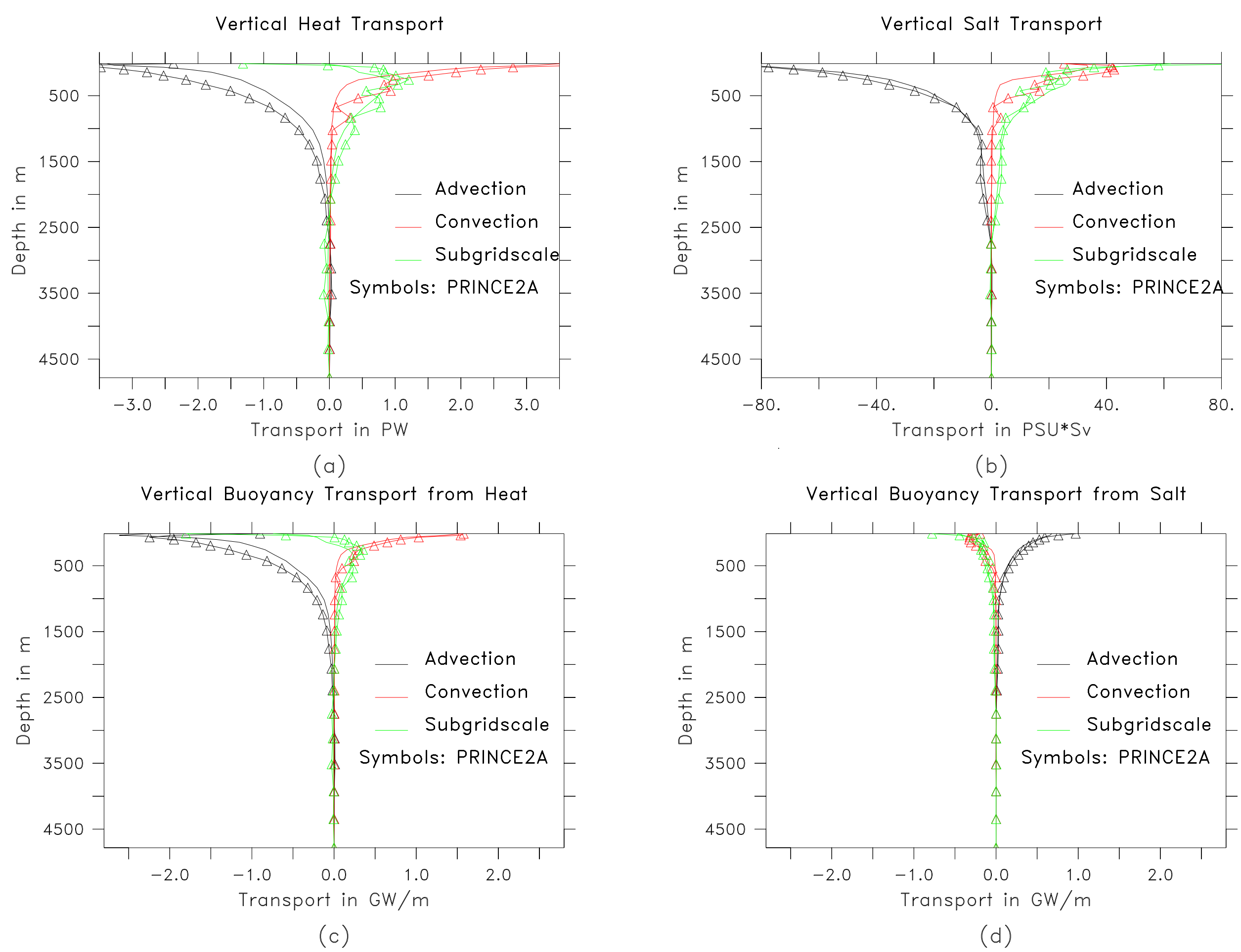

Figure 6: (a) Vertical heat fluxes due to advection, subgridscale mixing, and convection in the two models shown in Figure 1. (b) Vertical salt fluxes due to advection, subgridscale mixing, and convection. (c) Estimated buoyancy fluxes resulting from the heat fluxes in (a) (nonlinearity of equation of state means that results are not exact). (d) Estimated buoyancy fluxes resulting from the salt fluxes in (b).

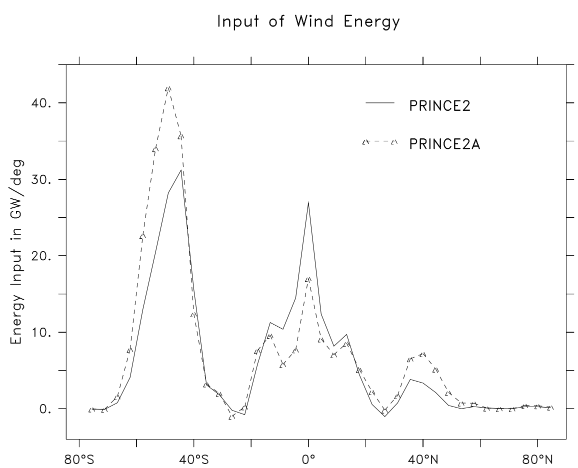

Figure 7: Wind stress input in the two models. Solid is PRINCE2, Dashed PRINCE2A.

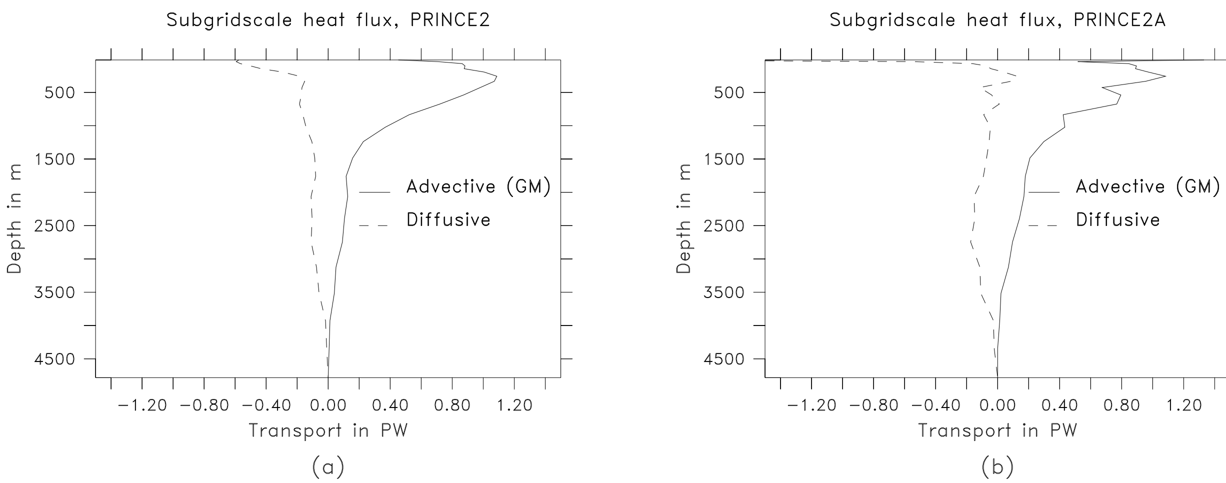

Figure 8: (a) Decomposition of the subgridscale heat flux into the component due to eddy-induced advection (solid) and diffusion (dashed) for model PRINCE2. (b) Same for model PRINCE2A.